Hydrological modeling of the Martian crust with application to the pressurization of aquifers

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[1] We develop a hydrological model of the Martian crust, including both ancient heavily cratered terrains and younger basaltic and sedimentary terrains. The porosity, permeability, and compressibility are represented as interdependent functions of the effective stress state of the aquifer, as determined by the combination of the lithostatic pressure and the fluid pore pressure. In the megaregolith aquifer model, the crust is modeled as a 2 km thick megaregolith, composed of lithified and fractured impact ejecta, overlying the impact-fractured and partially brecciated basement rock. The hydraulic properties depend primarily upon the abundance of breccia and the compressional state of the fractures. The porosity ranges from approximately 0.16 at the surface to 0.04 at a depth of 10 km, with a sharp discontinuity at the base of the regolith. The permeability varies from approximately $10^{-11}$ m$^2$ at the surface to $10^{-15}$ m$^2$ at depths of 5 km or more and is strongly dependent upon the fluid pore pressure. The hydrologic properties of basaltic and sedimentary aquifers are also considered. These parameters are used to model the fluid pressures generated beneath a thickening cryosphere during a postulated dramatic cooling of the climate at the end of the Noachian. As a result of a negative feedback between the fluid pore pressure and the permeability, it is more difficult than previously thought to generate pore pressures in excess of the lithostatic pressure by this mechanism. The production of the outflow channels as the result of such a climatic change is deemed unlikely.


1. Introduction

[2] There is a substantial body of evidence pointing to the fact that groundwater plays a central role in the Martian hydrologic cycle, both past and present. The dendritic valley networks, which are the earliest recorded fluvial features on Mars, were likely formed by some combination of surface runoff and groundwater sapping [Goldspiel and Squyres, 2000; Hynek and Phillips, 2003; Williams and Phillips, 2001]. Either process is suggestive of the existence of a widespread and shallow groundwater system during the Noachian epoch. More recently in Mars history, with ages ranging from Hesperian to Amazonian [Carr and Clow, 1981], the circum-Chryse outflow channels provide the strongest evidence for the importance of groundwater. These enormous fluvial features, ranging from tens to hundreds of kilometers in width and a kilometer or more in depth, originate from groundwater sources in chaos or canyon regions [Baker and Milton, 1974; Carr, 1979]. A number of smaller outflow channels are found throughout the planet, including Athabasca and Mangala Valles. The young ages of the channel surface of Athabasca Valles indicates the persistence of outflow activity into the past 10 to 100 Ma [Berman and Hartmann, 2002]. The most recent fluvial features, with ages possibly younger than 10 Ma, are small-scale gullies, which occur on the steep slopes of crater walls, central peaks, and canyon walls in high latitudes [Malin and Edgett, 2000a]. While the origin of these features is still uncertain, one interpretation is that they result from the freezing of small, confined aquifers at depths of several hundred meters [Gaidos, 2001; Mellon and Phillips, 2001].

[3] Despite this dramatic evidence for both ancient and recent hydrologic activity on Mars, many first order questions regarding the distribution of groundwater and the hydrologic properties of the Martian crust remain largely unanswered. Yet any attempt to understand the nature and history of the action of water on the Martian surface must hinge upon the assumptions made regarding the hydrologic properties of the subsurface. This study presents a generalized hydrologic model of the Martian crust that allows for the modeling of a variety of processes over the wide range of conditions thought to have existed in the Martian subsurface, and then applies that model to gain insight into the origin of the outflow channel floods.

[4] Hydrologic phenomena on Mars range in spatial scale from gullies at the hundred meter to kilometer scale, to outflow channels at the scale of tens to thousands of kilometers. For smaller scale features, the local variability in the
crustal materials and hydrologic properties will be significant, and the range of parameter space that must be considered will span several orders of magnitude. However, on large spatial scales, the local heterogeneity will average out and the large-scale hydraulic parameters can be estimated with a greater degree of confidence. The model here developed is a representation of the large-scale average hydraulic properties of the Martian crust. To a first approximation, Martian crustal materials can be divided into three categories: (1) heavily cratered, Noachian-aged crust that predominates in the southern highlands; (2) younger basaltic material that likely makes up much of the crust in the Tharsis and Elysium volcanic provinces, and which is likely present in smaller proportions within the crust elsewhere on the planet; and (3) sedimentary deposits that appear to be most abundant within closed depressions [Malin and Edgett, 2000b]. Many terrains are best represented as combinations of the three types, such as lava flows with interbedded sediments and impact ejecta, and the northern plains, which are thought to be made up of an ancient heavily cratered basement [Frey et al., 2002] overlain by volcanic plains with a thin sedimentary cover [Head et al., 2002]. We model the properties of the heavily cratered Martian crust based upon the record of surface impacts and the observed properties of terrestrial and lunar impact craters, drawing also on the hydrologic properties of terrestrial and lunar analogs to Martian regolith material. The younger basaltic and sedimentary regions are modeled directly after terrestrial aquifers.

[5] This model differs from previous studies in that it includes the compressibility of the aquifer, and thus allows for the modeling of time-dependent flow. Furthermore, all of the hydraulic properties of interest are modeled interdependently using a self-consistent and physically realistic model. The parameter values can be calculated for any combination of lithostatic and fluid pore pressure. This flexibility is essential for modeling many of the hydraulic processes that are thought to have occurred on Mars involving deeply buried aquifers and both large and rapidly varying pore pressures.

[6] In the following sections we review two existing models of the hydrologic properties of the Martian crust [Clifford, 1993; MacKinnon and Tanaka, 1989], before presenting the model that is developed in this study. This hydrologic model is then used to test the theory that the large pore pressures necessary to form the outflow channels could have been generated beneath a downward propagating freezing front [Carr, 1979].

2. Governing Equations

[7] Before developing the hydrologic model of the Martian crust, we first review the relevant governing equations in order to introduce the physical properties of interest. Steady state flow within an aquifer is governed by Darcy’s Law:

\[ q = K \nabla h, \]

where \( q \) is the volumetric flux vector of water per unit area or Darcy velocity (m/s), \( K \) is the hydraulic conductivity (m/s), and \( h \) is the hydraulic head (m). The hydraulic conductivity describes the resistance to flow within the aquifer, and is determined by the intrinsic permeability of the aquifer \( k \) (m²; 1 darcy = \( 10^{-12} \) m²), the fluid dynamic viscosity \( \mu \) (Pa s), the acceleration of gravity \( g \), and the density of water \( \rho_w \):

\[ K = k \frac{\rho_w g}{\mu}. \]

(2)

The dynamic viscosity of water is temperature dependent [Demming, 2002], described by the equation:

\[ \mu(T) = 2.4 \times 10^{-5} \cdot 10^{248/(T - 140K)} \text{(Pa} \cdot \text{s}). \]

(3)

For Martian gravity, the permeability can be converted to hydraulic conductivity values by multiplying by a factor of approximately \( 2 \times 10^8 \) (m s)⁻¹ at 273 K. In this study we report values of the permeability rather than the hydraulic conductivity, as the permeability is an intrinsic property of the aquifer material itself and does not require scaling to account for changes in gravity, fluid density, and viscosity. The hydraulic head is a potential term, which includes both the elevation \( z \) of a fluid parcel above a datum, as well as the aquifer pore pressure \( P_{pore} \):

\[ h = \frac{P_{pore}}{\rho_w g} + z. \]

(4)

Since we are only interested in the gradient of the head, it does not matter to what datum the elevation is referenced, as long as it is done consistently throughout the aquifer.

[8] Transient flow within an aquifer is governed by the consolidation equation [Domenico and Schwartz, 1990]:

\[ \frac{\partial h}{\partial t} = \frac{1}{S_s} \nabla \cdot (K \nabla h). \]

(5)

The specific storage, \( S_s \) (m⁻¹), describes the elastic response of the aquifer to pressure changes:

\[ S_s = \frac{\rho_w g}{S} (n^3 - 1). \]

(6)

where \( n \) is the porosity and \( S \) is the compressibility of water and the aquifer matrix respectively. The aquifer matrix compressibility is defined as

\[ \beta_{aquifer} = \frac{1}{V_{total}} \frac{\partial V_{total}}{\partial \sigma_{eff}} \approx \frac{1}{V_{pore}} \frac{\partial V_{pore}}{\partial \sigma_{eff}} = \frac{1}{1 - n} \frac{\partial n}{\partial \sigma_{eff}}. \]

(7)

Note that the compression of the aquifer depends upon the effective stress state \( \sigma_{eff} \), defined as the difference between the lithostatic and fluid pore pressures, rather than on the lithostatic pressure alone. Since the compressibility of the pore space is much greater than the compressibility of the actual mineral grains, the change in the volume of the rock, \( V_{total} \), with changing effective stress is essentially equal to the change in volume of the pore space, \( V_{pore} \). The final equality in equation (7) can be arrived at by setting \( n \) equal to \( V_{pore}/V_{total} \), and \( \partial V_{total} \) equal to \( \partial V_{pore} \) [Domenico and Schwartz, 1990].

[9] The porosity of the aquifer can be calculated as a function of the effective stress, again assuming that the
mineral grains themselves are essentially incompressible and the bulk compression is accommodated entirely by a decrease in the pore volume:

\[
\eta_{\text{eff}} = \left(\eta_0 - 1\right) \cdot \exp\left(\int_{0}^{z_0} \beta_{\text{aquifer}} \cdot d\sigma\right) + 1
\]

where \(\eta_0\) is the uncompressed porosity. As will be seen in the sections that follow, in some cases the compressibility is a function of the effective stress, and equation (8) must remain an integral expression.

[10] Thus three essential hydrologic parameters are required to represent an aquifer: the porosity, permeability, and compressibility. The porosity is important for estimating the total volume of water that may be present (section 4.5), for modeling the thermal conductivity of the crust (section 4.5), and for modeling pressurization by the injection of excess water into the pore space (section 6.2). The permeability, \(k\), is essential for modeling both steady state and transient flow within the aquifer (section 6.2). The compressibility, \(\beta\), governs both transient flow and the pressurization of the aquifer (section 6.2). These parameters are interdependent, as both the nature and the volume of the pore space determines the permeability, and the compressibility determines dependence of the porosity and permeability on the effective stress. Furthermore, as it is likely that extremely large pore pressures were required to form the outflow channels, it is necessary to consider the variation of these parameters both with depth and with changing pore pressure.

3. Previous Studies

[11] MacKinnon and Tanaka [1989] modeled the hydrologic properties of the heavily cratered Martian upper crust, consisting of a thick megaregolith composed of breccia overlying the impact-fractured basement rock. They assumed that the Martian crust is made up of initially impermeable igneous rock that has been fractured by impacts. Citing fracture widths resulting from the Danny Boy nuclear test explosion [Nugent and Banks, 1966], they represented the deep Martian crust as a basement rock permeated by fractures with an average aperture of 5 mm and a spacing of 3 m, resulting in a porosity of \(1.5 \times 10^{-3}\) and a permeability of approximately \(10^{-9}\) m\(^2\). They suggested that overlying this fractured basement rock is a 1–2 km thick megaregolith of impact ejecta, this based on Monte Carlo simulations of the ejecta thickness on a surface with a random distribution of impacts [Woronow, 1988] and on observations of apparent crustal discontinuities in this depth range. The hydraulic properties of this ejecta layer were modeled using clast size distributions from the Nevada Test Site [O’Keefe and Ahrens, 1985] and from the Meteor Crater ejecta. They calculated that this loosely packed Martian ejecta layer has a porosity between 0.1 and 0.2 and that the permeability is likely less than \(10^{-14}\) m\(^2\).

[12] While the MacKinnon and Tanaka [1989] model is a good conceptual representation of the ancient Martian crust, it does not include the variation of porosity and permeability with increasing confining pressure or changing pore pressure. The permeability of terrestrial materials decreases by several orders of magnitude at depths of a few kilometers [Huenges et al., 1997; Manning and Ingebritsen, 1999]. While fractures can have arbitrarily large apertures with little to no overburden or if they have been filled with particulate matter, the aperture of unfilled fractures decreases with increasing confining pressure. The fracture apertures measured at the site of the Danny Boy explosion are much larger than those typically found in the terrestrial crust, and are more similar to those measured in recently activated faults. Active motion along a fault is followed by a period of dramatically increased permeability due to the formation of large aperture fractures within the fault. However, the increase in aperture and permeability is a temporary effect and the wide apertures are not stable over geologic time [Gudmundsson, 2001].

This estimate of the fracture aperture may be representative of the time period immediately following an impact, but it is inflated above typical values for stable fractures at shallow depth by a factor of about 50, and neglects the variation of aperture with depth [Snow, 1970]. Since the permeability depends on the cube of fracture aperture, this overestimation of fracture aperture leads to overestimation of the permeability by a factor of as much as \(10^3\). Furthermore, their model does not include the compressibility of the aquifer and thus cannot be used to model aquifer pressurization or transient flows.

[13] The most widely cited model of the hydrologic properties of the Martian crust is that of Clifford [1981, 1993] and Clifford and Parker [2001]. This model assumed an exponential decrease in porosity with depth due to the elastic compression of the pore space and was based on the lunar model of Binder and Lange [1980] scaled to Mars gravity:

\[
\eta(z) = 0.2 \cdot e^{-z/6.5\text{ km}} \quad \text{Moon}
\]

\[
= 0.2 \cdot e^{-z/2.8\text{ km}} \quad \text{Mars.}
\]

where \(z\) is the depth in km below the surface. The Binder and Lange [1980] model was loosely based on lunar seismic velocities, which were originally interpreted to suggest closure of fractures and pores at the 20-km seismic discontinuity [Keihm and Langseth, 1977]. The lunar equation above simply fits an exponential decrease in porosity from 0.2 at the surface to 0.01 at 20 km depth, and does not attempt to fit the seismic data in between these values, which shows a discontinuity at the base of the megaregolith as well as the second discontinuity at 20 km [Toksoz et al., 1974; Watkins and Kovach, 1973].

[14] Alternatively, Toksoz et al. [1974] suggested that the 20 km seismic discontinuity may correspond to a compositional change rather than the closure of the pore space. Support for this interpretation of the seismic data came from analysis of the geoid to topography ratios around lunar basins, which indicates that the discontinuity is due to a density interface as would exist between an upper anorthositic crust and a lower noritic crust [Wieczorek and Phillips, 1997].

[15] More recent analyses of the lunar seismic data [Khan and Mosegaard, 2002; Khan et al., 2000; Longomne et al., 2003] have called into question the existence of the 20 km seismic discontinuity altogether. Khan et al. [2000], using
The demonstration of substantial crustal permeability at depths of 9 km on the Earth [Huenges et al., 1997], corresponding by a simple pressure scaling to a depth of 24 km on Mars and 56 km on the Moon, contradicts the interpretation of the elastic closure of the lunar pore space at 24 km on Mars and 56 km on the Moon, contradicting by a simple pressure scaling to a depth of 24 km on Mars and 56 km on the Moon, contradicting the matter further when both pores and fractures are present.

The proposed elastic decrease in porosity with depth does not correlate with the low compressibility actually measured for the lunar breccias [Warren and Trice, 1975]. The demonstration of substantial crustal permeability at depths of 9 km on the Earth [Huenges et al., 1997], corresponding by a simple pressure scaling to a depth of 24 km on Mars and 56 km on the Moon, contradicts the interpretation of the elastic closure of the lunar pore space at 20 km. In section 4.5, we demonstrate that the closure of pore space is more likely caused by either plastic deformation or pressure solution.

The Clifford [1993] study did not predict the permeability based on the model of porosity, since there is no unique relationship between porosity and permeability in a fractured porous medium. Clifford and Parker [2001] modeled the permeability of the Martian crust after the study by Manning and Ingebritsen [1999] of the average permeability in the terrestrial crust as a function of depth, which they again scaled to Mars gravity:

\[
\log(k) = -14 - 3.2 \log(z \cdot g_{\text{Mars}}/g_{\text{Earth}}) \\
= -12.65 - 3.2 \log(z),
\]

where \(z\) is the depth in km, and \(k\) is the permeability in m². The Manning and Ingebritsen [1999] model was based on indirect evidence for the permeability obtained from heat flow data and metamorphic reaction rates. [18] The permeability in the Earth’s crust is influenced by a number of different factors, including rock type, fracture frequency, fracture aperture, heat flow, degree of metamorphism, age, and tectonic history. A simple gravitational scaling of the terrestrial permeability function to Mars is likely an oversimplification. For the same reason, this model also does not allow for the variation of permeability with changing pore pressure. Note that as the depth approaches zero, or alternatively as the pore pressure approaches the lithostatic pressure, equation (10) would predict infinite permeability. This unphysical result occurs simply because the model was intended only to apply to greater depths within the Earth’s crust. Thus the Manning and Ingebritsen [1999] model is of only limited applicability and does not allow for the modeling of shallow aquifers or deep aquifers with large pore pressures, both of which are integral parts of the Martian hydrologic system.

Another drawback of these models is that the parameters of interest were not modeled self consistently. The porosity and permeability models were based on hydrologic models of different geologic environments on different planets. Furthermore, they did not consider the compressibility of the crust, thus precluding the modeling of time-dependant flows. What is needed is a single self-consistent and physically realistic model of the porosity, permeability, and compressibility of the Martian crust for any combination of depth and pore pressure.

4. Megaregolith Aquifer Model

4.1. General Characteristics

We assume, as a starting condition, that the crust of Mars at the time of its formation was composed of a crystalline basement rock [Harper et al., 1995], which was subsequently modified by impacts (Figure 1). On the basis of the effect of a single impact on the porosity and fracture frequency induced in the host rock, it should be possible to calculate the hydraulic properties of a planetary surface with a given crater distribution. Since the southern highlands of Mars are near crater-saturation, any preferential radial or circumferential orientation of the impact-induced damage should average out and the resulting porosity, permeability and compressibility are considered to be isotropic. We assume that the impacts during and after the period of heavy bombardment resulted in a heavily fractured and partially brecciated basement over lain by a megaregolith breccia layer of consolidated impact ejecta [Clifford, 1981; MacKinnon and Tanaka, 1989; Woronow, 1988]. For Noachian-aged surfaces, the thickness of this upper megaregolith layer would likely be between 1 and 3 km [Hartmann et al., 2001; MacKinnon and Tanaka, 1989; Ward, 2002; Woronow, 1988]. Since cratering and megaregolith production are contemporaneous processes, the megaregolith is subject to impact modification as well. This megaregolith layer is capped by a thin (tens of m up to 1 km) layer of aeolian and aqueous sediments [Malin and Edgett, 2000b], Viking-type “soil” [Moore et al., 1977], and unconsolidated regolith from the perpetual impact gardening [Hartmann et al., 1999].
2001; Shkaratov and Bondarenko, 2001; Watkins and Kovach, 1973). However, this upper thin veneer of materials will not be hydrologically active under the cold climate conditions that have persisted for much of Mars’ history.

4.2. Impact-Induced Breccias

[21] As impact-generated breccias are thought to be a significant component of the ancient heavily cratered portions of the Martian crust, it is thus necessary to estimate the distribution, abundance, and hydraulic properties of breccias in Martian aquifers. As previously stated, the accumulation of impact ejecta on the Martian surface likely produced a megaregolith breccia layer between 1 and 3 km thick. Furthermore, breccias are produced in the bedrock beneath craters both during the initial passage of the shockwave after impact, and during the subsequent decompression and modification phases [Dressler et al., 1996]. In this latter stage, large breccia bodies and anastamozing breccia dykes form as blocks of bedrock slide past one another during the modification of the transient cavity, providing a mechanism for the acoustic fluidization of the target material [Ivanov, 2002; Melosh and Ivanov, 1999].

[22] The distribution and abundance of breccia beneath impact craters can be inferred from a combination of gravity studies and drill core data. Early studies of terrestrial impact craters revealed that they are often associated with negative gravity anomalies [Innes, 1961; Pilkington and Grieve, 1992]. In a study of young lunar craters, Dvorak and Phillips [1977] found that the gravity data was best explained by a brecciated layer extending to a depth of one third of the crater diameter and laterally to the crater rim, with a density contrast of 0.3 g/cm³. The deep drilling project into the 40 km diameter Puchezh-Katunki impact crater found that the crust beneath the crater consists of blocks of bedrock, 100 to 400 m across, separated by zones of breccia [Ivanov et al., 1996; Kocharyan et al., 1996]. The average ratio of breccia to competent rock was 1:3 down to a depth of about 4 km. Combining these two lines of evidence, we suggest that the crust beneath large craters is composed of a mixture of breccia and bedrock, in a ratio of 1:3, down to a depth of 1/3 the crater diameter.

[23] The hydraulic properties of breccias can be estimated on the basis of observations of terrestrial and lunar breccias. Porous fault breccias have porosities of around 0.1 [Antonellini and Mollema, 2000]. MacKinnon and Tanaka [1989] calculated that Martian ejecta should have porosities in the range of 0.1 to 0.2. The porosities of lunar breccia samples lie in the range of ~0 to 0.4 with a mean of 0.17, and with the most representative sample of highlands breccia having a porosity of 0.15 [Warren and Rasmussen, 1987]. We adopt a value of 0.15 as representative of the uncompressed porosity of breccias, including both lithified impact ejecta and breccias intermixed with the bedrock at greater depth.

[24] Laboratory studies of the compressibility of fault breccia yield values of approximately 10\(^{-9}\) Pa\(^{-1}\) [Seront et al., 1982]. Similarly, the compressibility of lunar breccias is approximately 10\(^{-9}\) Pa\(^{-1}\) at low effective stresses, while at higher effective stresses it can be fit by a power law with a change in slope at 10 MPa [Warren and Trice, 1975]:

\[
\begin{align*}
\frac{1}{\text{B}_{\text{res}}(P)} &= c_1 \cdot \sigma_{\text{eff}}^2 \\
&= 7.94 \times 10^{-10} \quad \text{for } 0.1 < \sigma_{\text{eff}} < 10 \text{ MPa} \\
&= 2.81 \times 10^{-9} \quad \text{for } 10 \leq \sigma_{\text{eff}} \leq 1000 \text{ MPa} \\
&= -0.1 \quad \text{for } 0.1 < \sigma_{\text{eff}} < 10 \text{ MPa} \\
&= -0.65 \quad \text{for } 10 \leq \sigma_{\text{eff}} \leq 1000 \text{ MPa},
\end{align*}
\]

where the effective stress is in MPa and the compressibility is in Pa\(^{-1}\). The breccia porosity as a function of the effective stress can then be calculated using equations (8) and (11).

[25] Breccias have low permeability relative to their porosity due to the poor degree of sorting and the presence of small clast sizes. Gudmundsson [2001] measured the permeability of fault breccias to be in the range of 10\(^{-17}\) to 10\(^{-20}\) m\(^2\), much lower than the upper limit of 10\(^{-14}\) m\(^2\) calculated by MacKinnon and Tanaka [1989]. As will be seen in the next section, the permeability of intact breccia is orders of magnitude lower than the permeability of the superimposed fracture network, and thus can be neglected. Both seismic observations of the lunar megaregolith [Warren and Kovach, 1973], as well as geomorphic observations of fault scars in the top kilometer or so of the Martian megaregolith suggest that the breccia is likely a coherent, lithified unit. We assume that breccia is subject to impact fracturing to the same degree as competent rock, so fractured breccia is treated in the same manner as fractured bedrock with regards to the permeability. The observation of hydraulically conducting fractures within the breccia filled cores of fault zones [Gudmundsson et al., 2001] supports this assumption.

[26] In summary, we assume that the heavily cratered Noachian-aged crust is composed of a 2 km thick megaregolith overlying the partially brecciated basement rock, in which the ratio of breccia to bedrock is 1:3. The breccias in the upper megaregolith and intermixed within the bedrock are assumed to have similar hydraulic properties, with an uncompressed porosity of approximately 0.15, a compressibility given by equation (11), and negligible permeability. The permeability within the breccia units will be determined by the presence of the superimposed fracture network, if one exists.

4.3. Impact-Induced Fractures

[27] The permeability resulting from the presence of fractures beneath an impact crater is determined by the fracture frequency (N, expressed in terms of the number of fractures per unit distance) and the fracture aperture (b), as described by the cubic law [Domenico and Schwartz, 1990]:

\[
k_{\text{frac}} = \frac{N b^3}{12}.
\]  

The fracture porosity in the host rock can be calculated as the product of the fracture frequency and the fracture aperture:

\[
\phi_{\text{frac}} = N \cdot b.
\]

Thus, to calculate the impact-induced permeability and fracture porosity, it is necessary to estimate the frequency and aperture of fractures beneath a crater.
Seismic data indicate that the zone of fractures around a crater with diameter $D$ extends radially a distance of $D/2$ beyond the crater rim and down to a depth of $D/2$ [Ackermann et al., 1975]. This is supported by both impact experiments [Ahrens and Rubin, 1993] and numerical modeling studies [Ivanov, 2002]. The fracture frequency in the 218 m deep drill hole in the Lappajarvi crater in Finland [Kukkonen et al., 1992] is approximately 20 m$^{-1}$. Similarly, the fracture frequency in outcrops of the intensely fractured and shattered bedrock in the Acraman crater in Australia is between 20 and 100 m$^{-1}$ [Williams, 1994]. These values are significantly greater than the fracture frequency of typical terrestrial crust, which lies in the range of 0.1 to 5 m$^{-1}$ [Seeburger and Zoback, 1982], but are roughly consistent with the fracture frequency of about 10 m$^{-1}$ in active fault zones [Gudmundsson et al., 2001]. The increased fracture frequency is in agreement with the observation of a large zone of crushed rock in the vicinity of underground nuclear explosions in which the density of joints increases up to hundreds of times above the initial value [Kocharyan et al., 1996]. For this work we adopt a fracture frequency of 20 m$^{-1}$.

A number of both theoretical and experimental studies have been performed on the compression of fractures. Experimental studies measure the change in aperture and permeability with changing confining pressure for small samples of actual fractures [Cook, 1992]. Theoretical studies, on the other hand, produce complicated integral formulations for the aperture as a function of pressure, based on the statistical distribution of asperities along the fracture faces [Brown and Scholz, 1986]. The theoretical studies base the geometry of the fracture face on measurements of small samples of actual fractures, thus both the experimental and theoretical approaches rely on laboratory-scale fracture samples. However, it has been demonstrated that the permeability of the Earth’s crust is very strongly scale dependent [Brace, 1984]. Permeability on the scale of a laboratory sample is often orders of magnitude less than the large-scale permeability in the same region. Furthermore, it has been demonstrated that flow within fractures is focused along discrete channels between the fracture faces [Wang et al., 1988]. Hence the fractal-like nature of the fracture face topography and the nonuniformity of the flow within fractures make it unlikely that the hydraulic behavior of small-scale laboratory samples of natural fractures will be representative of the average behavior of large-scale in situ fractures.

We model fracture apertures using a novel approach in which we use data on near-surface fracture apertures to calculate the fracture aperture and compressibility at low effective stresses. We then extrapolate these values to high effective stresses using data on the permeability and fracture frequency at greater depths.

Snow [1970] calculated the effective fracture apertures in the near-surface on the basis of permeability and fracture frequency measurements in wells to depths in excess of 100 m in a variety of rock types, and found that the apertures decrease regularly with depth and are essentially independent of host rock type. We plot these data together on one graph (Figure 2), and fit the data with an exponential function:

$$b(z)_{\text{Earth}} \approx 2 \times 10^{-4} (m) \cdot e^{-0.0087(m^{-1})z} \quad (14)$$

where $b$ and $z$ are both expressed in meters. The exponential decrease in aperture implies a constant compressibility of the fractures in the near surface. The fracture compressibility is then calculated as

$$\beta_{\text{frac,0}} = \frac{1}{V_{\text{frac}}} \frac{dV_{\text{frac}}}{d\sigma_{\text{eff}}} = \frac{1}{\rho g \cdot b(z)} \frac{db(z)}{dz} \approx 2.9 \times 10^{-7} \text{Pa}^{-1}$$

$$\beta_{\text{rock}} = \frac{1}{V_{\text{rock}}} \frac{dV_{\text{rock}}}{d\sigma_{\text{eff}}} = \frac{1}{V_{\text{rock}}} \frac{dV_{\text{frac}}}{d\sigma_{\text{eff}}} = \beta_{\text{frac,0}} \cdot N_{\text{frac,0}} \cdot b_{\text{frac}} \quad (15)$$

where $\beta_{\text{frac,0}}$ is the compressibility of an individual fracture under low effective stresses, $\beta_{\text{rock}}$ is the whole rock compressibility due to the presence of the fractures, $\rho$ is the rock density, and $V_{\text{frac}}$ is the volume of the fracture.

We estimate fracture apertures at depths greater than 100 m on the Earth based on permeability data by assuming a typical fracture frequency. Seeburger and Zoback [1982] measured fracture frequencies to depths in excess of 1 km in three geologically distinct areas. On the basis of their data, we consider a range of 0.1 to 1 m$^{-1}$ as being representative of the fracture frequency of the Earth’s crust at depths of a kilometer or more. Permeabilities obtained in the KTB drill hole of the German Continental Deep Drilling Program are on the order of $10^{-16}$ to $10^{-17}$ m$^2$ for depths ranging from 0.5 to 5 km [Huenges et al., 1997]. Using this range in permeabilities with the above range in fracture frequencies, equation (12) allows us to calculate the typical fracture aperture at large depths to be on the order of $10^{-5}$ to $5 \times 10^{-6}$ m. These minimum effective apertures are in agreement with laboratory experiments on the compression of fractures [Cook, 1992]. While an aperture of 10 μm seems exceptionally small to be considered an open fracture, it must be remembered that this is an effective hydraulic aperture. The actual fracture would have a larger aperture in places and be completely closed at the contact points between the two faces.

Figure 2. Fracture aperture as a function of depth for a variety of rock types (data of Snow [1970]).
Fracture compressibility must decrease markedly with depth, approaching zero so as to agree with the relatively constant permeability at large depths found by Huenges et al. [1997]. This can be justified conceptually by considering that on the microscopic level fracture surfaces have a fractal-like roughness [Wang et al., 1988]. Thus the aperture is just an average value, while the fracture is held open by the asperities on the fracture surface. As the fracture is compressed, the opposing faces come into contact in more places and the resistance to further compression is increased proportionately (Figure 3). We apply this conceptual model by assuming a simple exponential decrease in the fracture compressibility as a function of confining pressure, and fitting it to the surface compressibility and the inferred minimum fracture aperture at large depths:

$$\beta_{frac}(\sigma_{eff}) = \beta_{frac,0} \exp\left(-\frac{\sigma_{eff}}{\sigma_{frac,0}}\right)$$

$$b_{frac}(\sigma_{eff}) = b_0 \cdot \exp\left[-\int_0^{\sigma_{eff}} \beta_{frac}(\sigma_{eff}) \cdot d\sigma\right]$$  \hspace{1cm} (16)

$$= b_0 \cdot \exp\left[\beta_{frac,0} \beta_{frac, \sigma_{frac,0}} \left(e^{-\sigma_{frac,0}/\sigma_{frac,0}} - 1\right)\right],$$

where $\sigma_{frac,0}$ is the scale stress of the exponential decrease in fracture compressibility. Assuming a minimum fracture aperture of 10^{-5} m at large depth, we find that $\sigma_{frac,0}$ is approximately 10.3 MPa. This aperture function fits both the near surface aperture data as well as the deep permeability data. The general character of the plot of the fracture aperture as a function of effective stress in this study (Figure 4) agrees well with both the theoretical and laboratory studies, while avoiding the limitations of small-scale samples. This approach, while more empirical in nature, is likely a better representation of the net behavior of real fractures within the crust.

[14] In summary, we assume that heavily cratered crust has a fracture frequency of approximately 20 m^{-2}. The fracture apertures decrease exponentially with increasing effective stress at low values of the effective stress, from an uncompressed value of approximately 2 × 10^{-4} m. Under conditions of high effective stress, the fracture apertures plateau at a value of approximately 10^{-5} m, while the fracture compressibility decreases to zero. The fracture aperture and compressibility as expressed in equation (16) depend only upon the effective stress state of the aquifer, making these results directly applicable to Mars.

4.4. Megaregolith Aquifer Model Synthesis

[33] We use the above description of the hydrologic properties of impact-induced breccias and fractures to synthesize an aquifer model based on the record of impacts on the surface. As stated earlier, we assume that the top 2 km of the crust is composed of a lithified and fractured breccia. Beneath this megaregolith, the crust is composed of an equally fractured mixture of bedrock and breccia. We assume that the effects of superimposed impacts do not add, as preexisting fractures and faults would likely be reactivated by subsequent impacts. Thus we assume that the fracture frequency and breccia distribution beneath the crater-saturated Noachian-aged crust is the same as that found beneath a single isolated crater. Since the largest fraction by area of the crater population on Noachian aged surfaces is composed of craters with radii in excess of 15 km, there will likely be no significant decrease in the fracture frequency and breccia abundance in the top 15 km of the crust. As will be seen in section 4.5, this likely constitutes the bulk of the hydrologically active crust of Mars. Crater saturation will also lead to an averaging out of any preferred radial and circumferential orientations of the impact-induced damage beneath individual craters. It is thus assumed that crater-saturated terrains are underlain by a uniform and isotropic distribution of fractures and breccia to a depth of 15 km or more, with a breccia fraction of 0.25 and a fracture frequency of 20 m^{-1}. The hydraulic properties of less densely cratered terrains could be calculated using the crater density, though at low crater densities the assumptions of uniform and isotropic hydraulic parameters would break down.

[36] The porosity in the crust is largely a result of the presence of breccias, but it also includes smaller components due to the fracture porosity and the primary porosity of the crystalline rocks:

$$n_{total}(z, \sigma_{eff}) = f_{brecc}(z) \cdot \left[n_{brecc} - 1\right] \cdot \exp\left[\int_{\sigma_{0}}^{\sigma_{eff}} b_{frac}(\sigma_{eff}) \cdot d\sigma\right] + 1] + N_{frac,bbrecc}\left[+ f_{brecc}(z) + \left(n_{primary} - 1\right) \cdot \exp\left[\beta_{primary} \cdot \sigma_{eff} \right] + 1\right],$$  \hspace{1cm} (17)

where $f_{brecc}$ is the breccia fraction within the crust as a function of depth (defined as 1.0 in the megaregolith, and 0.25 in the partially brecciated bedrock at greater depths), $n_{primary}$ is the unbrecciated bedrock primary porosity (assumed to be 0.02 [Domenico and Schwartz, 1990]), $\beta_{primary}$ is the compressibility of the primary pore space in the bedrock (assumed to be 10^{-10} Pa^{-1} [Domenico and Schwartz, 1990]), and all other parameters are as previously defined. The permeability is determined by the fracture frequency and aperture as described by equations (12) and (16). The net compressibility of the crust depends on the fracture compressibility, the fracture frequency, the compressibility of the breccia, the volumetric breccia fraction in the crust, the primary pore space compressibility in the rock,
and the compressibility of the water within the pore space ($\beta_{\text{water}}$):

$$\beta_{\text{total}} = \beta_{\text{frac}} \cdot N_{\text{frac}} \cdot f_{\text{frac}} + f_{\text{brecc}} \cdot \beta_{\text{brecc}} + (1 - f_{\text{brecc}}) \cdot \beta_{\text{primary}} + n \cdot \beta_{\text{water}}.$$  (18)

A summary of all parameters used in the equations is presented in Table 1.

[37] The megaregolith aquifer model of this study is presented in Figure 4 alongside the model of Clifford [1993] and Clifford and Parker [2001], as well as the sandstone model to be developed in section 6. The parameters for the megaregolith model in Figure 4 are calculated on the basis of the assumption of hydrostatic pore pressure, though similar plots can be produced for any amount of excess pore pressure.

[38] In an attempt to quantify the uncertainty in the megaregolith hydrologic model, we calculate the likely range of parameter space using a Monte Carlo analysis. We consider the likely range of values for the key input parameters for which the uncertainty can be estimated (Table 1). Note that for the fracture model, uncertainty is considered in the uncompressed fracture aperture and fracture frequency only. The fracture compressibility function arrived at above is constrained by the minimum fracture aperture at great depth, and thus is not independent of the assumed uncompressed aperture. The adopted range in uncompressed fracture aperture leads to the likely range in minimum fracture aperture discussed above without varying the fracture compressibility function. We represent the uncertainty in the breccia compressibility function expressed in equation (11) by a scaling factor ranging from 0.3 to 3, encompassing the range in the data of Warren and Trice [1975]. The range in values for each parameter is assumed to correspond to the two-sigma variation. These parameters are then varied normally using a random number generator, and the mean values and standard deviations are
4.5. Closure of Pore Space and the Base of the Aquifer

Previous studies [Clifford, 1981, 1993] emphasized the depth of the base of the aquifer as a means of calculating the total volume of fluid that could be contained within the pore space of the Martian crust, and thus constraining the total water storage capacity of the crust and the amount of water available for surface processes. Clifford [1993] assumes that it is the elastic closure of the pore space with increasing pressure that determines the base of the aquifer. However, using the actual compressibilities of breccias and fractures, we find that a significant amount of pore space and permeability extends to depths much greater than the 8 km depth assumed in that study. Considering elastic compression alone, the porosity decreases slowly with depth beyond the base of the megaregolith, and will remain at values of around 0.04 to 0.03 until a depth is reached at which the impacts into the surface had little effect. Similarly, fracture apertures reach a minimum value of approximately 10^{-3} m, and thus continue to be hydraulically conducting even under conditions of extremely large overburden pressures.

Instead, we propose that an upper limit to the thickness of aquifers on Mars is imposed by the rheology of the rocks. The brittle to ductile transition is defined as the depth at which the stress required to form a new fracture is equal to the frictional resistance on existing fractures [Byerlee, 1968]. The ductile flow in this case is still cataclastic in nature, and is not relevant to the closure of the pore space. The brittle to plastic transition (BPT), which is more relevant to the closure of pore space and fractures within an aquifer [Brace and Kohlstedt, 1980], occurs approximately when the differential stress necessary for plastic flow is equal to the effective confining pressure [Kohlstedt et al., 1995]. We calculate the maximum differential stress that could be supported in the Martian crust at the end of the heavy bombardment using a combined thermal and rheological model of the Martian crust based on the megaregolith model developed above.

The thermal conductivity of the Martian crust is calculated for both dry and wet conditions. For dry breccias, the thermal conductivity is given by

\[ k = 2.0 \cdot e^{-n/m} \left( W \cdot m^{-1}K^{-1} \right), \]

where \( k \) (W m^{-1}K^{-1}) is the thermal conductivity, \( n \) is the porosity, and \( n_0 \) is 0.0738 [Warren and Rasmussen, 1987]. For saturated breccias, the thermal conductivity is calculated by the harmonic mean of the thermal conductivities of rock and water:

\[ k = \frac{k_r \cdot k_w}{n \cdot k_r + (1-n) \cdot k_w}. \]

where \( k_r \) and \( k_w \) are the thermal conductivities of rock and water, taken to be 3 W m^{-1}K^{-1} and 0.57 W m^{-1}K^{-1} respectively [Demming, 2002]. In the partially brecciated basement crust, the thermal conductivity is likely intermediate between the harmonic and arithmetic means of the conductivities of the breccia component and solid rock. For the saturated crust, the harmonic and arithmetic means yield essentially the same result due to the similar thermal conductivities of rock and saturated breccia. At depths greater than the depth of the BPT, it is assumed that the pore space has closed and the thermal conductivity is set equal to that of solid rock.

The heat flux into the base of the lithosphere at the end of the heavy bombardment is taken to be 65 mW m^{-2} [Hauck and Phillips, 2002], and the surface temperature is set to be 273 K on the basis of the assumption of a warmer and wetter early Mars. If surface temperatures were cooler, as is suggested by some [e.g., Sfyries and Kasting, 1994], the brittle to plastic transition would be pushed deeper within the crust, increasing the thickness of potential Martian aquifers.

The rheology of the Martian crust is modeled after studies of the rheology of diabase in dislocation creep:

\[ \dot{\varepsilon} = A \cdot \sigma^n \exp(-E_a/RT), \]

where \( A, n, \) and \( E_a \) are constants that take on different values for dry [Mackwell et al., 1998] or wet [Caristan, 1982] diabase. For the case of a dry crust, it is clearly appropriate to use the rheology of dry diabase, but it is calculated for the fracture aperture, porosity, permeability and compressibility as a function of depth. Values of the permeability and compressibility are varied and averaged logarithmically. The one-sigma variations in the hydraulic parameters are represented in Figure 4 by the shaded gray regions.
unclear which rheology is more appropriate for the case of water saturated pore space. For pore pressures greater than the hydrostatic pressures, the water fugacity is likely to be large and the wet rheology may be appropriate. However, experiments on wet diabase experienced partial melting upon release of the water, which would have resulted in an overly weak rheology [Mackwell et al., 1998]. Thus the rheologies of both dry and wet diabase are plotted for the case of the saturated crust, while only the dry rheology is plotted for the case of unsaturated crust.

[44] Figure 5 shows that the BPT occurs between depths of 6 and 10 km for the dry crust case, and between depths of 17 and 26 km for the case of saturated crust. The 26 km depth is likely an overestimate, as the dry rheology assumed is probably stronger than the wet Martian crustal rocks. Furthermore, even for the case of a wet early Mars, an upper dry layer of regolith may have formed an insulating blanket on the surface, thus decreasing the depth to the BPT. Similarly the 6 km depth is likely an underestimate, as it is unlikely that no water existed in the crustal pore space on early Mars. The unrealistic assumption of a fully dry Martian crust results in excessively low thermal conductivities and high temperatures at shallow depth, leading to partial melting at depths as shallow as 30 km under Noachian conditions. The approximate range of likely depths for the BPT at the end of the heavy bombardment, and thus the upper limit for the thickness of ancient Martian aquifers, is between 10 and 20 km. However, in regions of continued geologic activity and beneath young impact craters it is possible that open pore space has been created at greater depths later in Mars history when the BPT was pushed deeper by the lower surface temperatures and decreased heat flux.

[45] While the brittle to plastic transition imposes an upper limit on the thickness of aquifers on Mars, other processes can act to close off the pore space at shallower depths. Terrestrial observations demonstrate the importance of pressure solution in the closure of both fractures and primary pore space in the presence of a fluid [Renard et al., 2000]. In short, this results from preferential dissolution at
the contact points of grains and fracture surfaces, followed by preferential precipitation in the void spaces. This is a time dependent process that is strongly sensitive to the temperature, pressure, rock composition, and the microscopic geometry of both the contacts and void spaces. Furthermore, fractures and fault breccias exhibit markedly different behaviors with regards to their susceptibility to pressure solution. The process is poorly constrained in terrestrial systems, and cannot be reliably applied to Mars. Additionally, hydrothermal circulation within the crust [Travis et al., 2003] could facilitate dissolution and mineral precipitation, closing off the pore space and emplacing veins within the fractures.

Claims regarding the total water storage capacity of the Martian crust will be limited by this uncertainty in the depth of closure of the pore space. We use the megaregolith model above to estimate the total pore volume of the Martian crust for a range of aquifer thicknesses. For aquifer thicknesses of 5, 10, and 20 km we calculate total pore volumes of 6.1, 9.8, and $17 \times 10^7$ km$^3$ respectively. The pore space present in basaltic aquifers within the Tharsis rise and the northern plains must also be considered. The total volume of Tharsis basalts has been estimated at $3 \times 10^8$ km$^3$ [Phillips et al., 2001]. Taking an average basal porosity of 0.1 (see section 5.1), this would give a total pore volume of $3 \times 10^7$ km$^3$. For a more conservative estimate, we also consider cases in which there is significant porosity only in the top 2 to 5 km of the crust in this region, as could result from closure of the pore space at depth or if a significant amount of low-porosity volcanic intrusions contributed to the Tharsis rise, leading to total pore volumes of 2.5 to $6.3 \times 10^6$ km$^3$. Representing the northern plains as a 1 km thick layer of porous basalt, we find a total pore volume of $4.8 \times 10^6$ km$^3$ above the underlying megaregolith. Using the above ranges in values, we get low and high estimates of the total pore volume of $6.8 \times 10^7$ and $2.0 \times 10^8$ km$^3$, with an intermediate estimate of $1.1 \times 10^8$ km$^3$. These values are in agreement with the estimates of Clifford [1981], despite the differences in the hydrologic model used. However, we again caution the reader with respect to the great uncertainties in these estimates.

5. Alternative Aquifer Types

5.1. Basaltic Aquifer Model

While the megaregolith aquifer model is likely representative of much of the Noachian-aged highlands, the younger surfaces that predominate in other areas of the planet have not been significantly modified by impacts. The Tharsis and Elysium volcanic provinces are composed of thick sequences of basalts flows, the upper portions of which are young enough to have escaped the heavy bombardment of the early Noachian era. The northern plains, as well as a number of other areas across the planet, are likely composed of layers of basaltic lava flows and sediments overlying more ancient cratered terrain beneath [Frey et al., 2002; Head et al., 2002].

Terrestrial basaltic aquifers are characterized by both high porosity and high permeability. While there is great variability in the hydraulic properties of terrestrial basalts, it is likely that the average properties of terrestrial and Martian basalts are similar. Over small spatial scales, it is necessary to model basaltic aquifers using a stochastic approach to represent the significant heterogeneity of the aquifers [Welhan et al., 2002]. However, over the large spatial scales commonly of interest on Mars, this lateral heterogeneity averages out and it is appropriate to use the bulk properties. [46] The hydraulic properties of the Snake River Plains aquifer summarized in studies by Gego et al. [2002] and Welhan et al. [2002] are reviewed here. A basaltic aquifer is made up of a thick layer of massive basalt (~84% by volume), overlain by a thinner interflow zone (~12% by volume) [Gego et al., 2002]. These interflow zones are both very porous and heavily fractured, as a result of thermal cooling of the outer surface of the flow and rotational stresses at the lava flow front. The interflow zones have an average porosity of 0.22 and an average permeability of approximately $10^{-8}$ to $10^{-10}$ m$^2$. The intervening layers of massive basalt, on the other hand, have an average porosity of 0.09 and permeability of approximately $10^{-11}$ to $10^{-13}$ m$^2$.

The average porosity of a basaltic aquifer is then calculated to be 0.10. The average horizontal and vertical permeabilities can be calculated using the arithmetic and harmonic means, yielding values of approximately $10^{-9}$ to $10^{-11}$ m$^2$ and $10^{-11}$ to $10^{-13}$ m$^2$ respectively. If there is sufficient time between individual eruptions, a layer of sediments will accumulate above each lava flow, which on Mars could include impact ejecta as well as aeolian and fluvial sediments. These sediments can act to further increase the porosity and permeability, or alternatively they may fill the fractures in the underlying interflow zone and result in a net decrease in the porosity and permeability.

There is not enough data to accurately constrain the behavior of basaltic aquifers under varying conditions of confining stress and pore fluid pressure, making the basaltic aquifer model of limited applicability. While some of the permeability is due to simple fractures, much of it is due to the presence of larger interconnected cavities within the interflow zones. The permeability is likely less sensitive to the effective stress than was found for the megaregolith model due to the presence of these larger cavities, though it is unlikely that the high values measured at the surface persist to depths of several kilometers as some compression is inevitable.

In a study of basaltic aquifers in the Oregon Cascades, Saar and Manga [2004] modeled the permeability as a function of depth based on a number of indirect approaches, including spring discharges, thermal profiles, magmatic intrusion rates, and seasonal seismicity. They represent the permeability as following an exponential decrease with depth in the top 800 m, followed by a power law relationship similar to that of Manning and Ingelbritsen [1999] at greater depths. The horizontal permeability ranges from $10^{-7}$ to $10^{-11}$ m$^2$ at the surface, to $10^{-15}$ to $10^{-18}$ m$^2$ at a depth of 10 km. They propose that the decrease in permeability with depth is exclusively due to mineral precipitation, though we would suggest that the elastic compression of the fractures plays a significant role as well. Manga [2004] suggests that lower water to rock ratios on Mars would lead to negligible mineral precipitation and the maintenance of the high surface permeability values even at great depth. However, as discussed earlier, pressure solution
and hydrothermal circulation are likely important in the closure of pore space and the decrease in permeability with depth on Mars. While the Saar and Manga [2004] study sheds some light on the variation of permeability with depth in basaltic aquifers, the methods of calculating the permeability were indirect and there are large uncertainties in the numbers. There still remains much work to be done to fully understand the dependence of the permeability, porosity, and compres- sibility on the depth and pore pressure in basaltic aquifers on both Earth and Mars.

5.2. Sedimentary Aquifers

A number of studies have emphasized the possibility of extensive sedimentary deposits on Mars [Goldspiel and Squyres, 1991; Malin and Edgett, 2000b]. Sedimentary rocks can display a wide range of hydraulic behaviors, and a comprehensive treatment of the topic is beyond the scope of the present study. We here aim only to outline the basic properties and behavior in such a way as to allow for the production of a generalized model. The porosity of sandstone at the surface is generally in the range of 0.05 to 0.3 [Domenico and Schwartz, 1990], and is commonly modeled as decreasing exponentially with depth [Athy, 1930; Chapman et al., 1984]:

$$ n(z) = n_0 e^{-z/z_0}, \quad (22) $$

where $n_0$ is the surface porosity and $z_0$ is the scale depth. Chapman et al. [1984] used values of 0.25 for $n_0$, and 3 km for $z_0$. It is significant that this exponential decrease in porosity cannot be attributed to the simple elastic compression of the pore space. Laboratory studies indicate that the porosity in sandstone decreases elastically much more slowly with increasing effective stress than would be predicted on the basis of the observed variation of the porosity with depth [Demming, 2002; Neuzil, 1986]. The rapid decrease in porosity with depth observed in sandstones is likely due to mineral precipitation within the pore space or perhaps to a long-term viscoelastic response, and is more dependent upon the thermal history than the effective stress [Neuzil, 1986]. The elastic response of the pore space to changes in the fluid pressure is much less significant than the variation of the porosity with depth, amounting to an increase in porosity of 0.015 with an increase in pore pressure of 10 MPa. As a model of sandstone aquifers on Mars, we adopt the formula of Chapman et al. [1984] and neglect the change in porosity with changing effective stress state. We do not scale the value of $z_0$ to Mars gravity, since the decrease in porosity with depth is not due to elastic compression. While it is unknown exactly what processes are responsible for the decrease in porosity with depth in terrestrial sandstones and how this would scale to Mars, the above relationship should capture the basic behavior of Martian sandstones. On the basis of the summary of Neuzil [1986], we represent the compressibility as an exponentially decreasing function of the effective stress:

$$ \beta(\sigma_{eff}) = \beta_{0,\text{sed}} \cdot \exp(-\sigma_{eff}/\sigma_{0,\text{sed}}), \quad (23) $$

where we assume values for $\beta_{0,\text{sed}}$ of approximately $10^{-9}$ Pa$^{-1}$, and $\sigma_{0,\text{sed}}$ of 25 MPa. The permeability of a sedimentary rock is dependent upon the porosity, grain size, and degree of sorting of the particles [Nelson, 1994]. It is commonly assumed that the permeability is proportional to the log of the porosity [Archie, 1950; Demming, 2002], but no one relationship between permeability and porosity can be said to be representative of all sandstones or shales. On the basis of the data of Nelson [1994], we represent the log of the permeability as being proportional to the porosity:

$$ \log_{10}(k) = -17.3 + 25 \cdot n, \quad (24) $$

where $k$ is the permeability in m$^2$, and $n$ is the porosity. The porosity, permeability, and compressibility of the sandstone aquifer model are represented in Figure 4 alongside the megaregolith aquifer model. An important difference between the sandstone and megaregolith aquifers is now apparent. In the megaregolith model the permeability is dependent on the elastic closure of the fractures and scales directly with the effective stress state, whereas in the sandstone model the permeability and thus the permeability are much less sensitive to the effective stress state.

5.3. Mixed-Media Aquifers

Single component aquifer models may be reasonable representations of certain regions of the Martian crust. For example the upper portion of the Tharsis region is composed largely of basalts, while the deep crust in the heavily cratered southern highlands can be represented by the megaregolith model. However, in many cases it may be necessary to consider the effects of a mixture of two components, such as sediments or basaltic lava flows interlayered within or overlying a megaregolith.

For the case of isotropic, intimate mixtures of different components, the permeabilities can be averaged using the geometric mean [Demming, 2002]:

$$ k_{eff} = \prod \left( k_i^f \right), \quad (25) $$

where $k_i$ and $f_i$ are the permeability and volumetric fraction of the $i$th component, respectively. Alternatively, for the case of horizontal layering of aquifer components, the permeability is anisotropic. The horizontal permeability is found by the arithmetic mean, while the vertical permeability is calculated using the harmonic mean:

$$ k_{hor} = \sum (f_i \cdot k_i) $$

$$ k_{vert} = \left[ \sum (f_i/k_i) \right]^{-1}. \quad (26) $$

The effective compressibility for mixed media aquifers can be found by a simple arithmetic mean of the compressibilities of the components.

6. Application: Aquifer Pressurization Through Climate Change

6.1. Background

The outflow channels are commonly thought to have formed when the fluid pore pressure within a confined aquifer reached or exceeded the lithostatic pressure [Carr, 1979]. Various studies ascribe these fluid pressures to be the...
result of (1) a perched aquifer [Carr, 1979, 1996], (2) a confined aquifer pressurized by a downward propagating freezing front [Carr, 1979], (3) artesian pressures caused by the proposed uplift of Tharsis [Clifford and Parker, 2001], (4) the intrusion of magma into ice rich crust [McKenzie and Nimmo, 1999; Squyres et al., 1987], (5) the decomposition of gas hydrates [Milton, 1974], and (6) the coalescence of groundwater flow in the tectonic fractures surrounding Tharsis [Baker et al., 1991]. While many of these ideas are conceptually valid, it has not been possible to adequately test them without a complete hydrologic model of the Martian crust.

We here apply our hydrologic model to investigate the possibility that the fluid pressures necessary to form the outflow channels could be generated beneath a downward propagating freezing front as the result of a rapid cooling of the climate, as proposed by Carr [1992]. As the aquifer progressively freezes, the volumetric expansion of water accompanying the phase change would pressurize the remaining aquifer beneath. The premise of this hypothesis relies on a dramatic cooling of the Martian climate to its current cold state. Mars’ climate evolution is still poorly understood, but there is clear evidence that the surface of Mars was both warmer and wetter during the Noachian epoch, including the high drainage densities of the valley networks [Baker, 1982; Hynek and Phillips, 2003] and the inferred high erosion rates [Carr, 1992]. The climatic evolution was likely driven, to a large degree, by the loss of an early thick greenhouse atmosphere to impact erosion, solar wind sputtering, adsorption of CO2 onto the regolith, and the possible precipitation of carbonates [Brain and Jakosky, 1998; Melosh and Vickery, 1989]. Alternatively, other workers propose that the Martian climate may have been subject to more rapid, episodic climate changes associated with redistribution of the volatile inventory [Baker et al., 1991].

For perfectly confined aquifers, arbitrarily large pressures could be generated by the growth of the cryosphere, potentially leading to catastrophic breakup and the formation of the outflow channels [Carr, 1979]. However, over large spatial and temporal scales, all aquifers are likely interconnected to some degree [Clifford, 1993]. This interconnectedness would be even more pronounced in the highly fractured megaregolith of the Martian southern highlands, as well as in the thick sequences of basalt in the Tharsis rise. Thus it seems likely that the entire circump-Chryse region was underlain by an extensive aquifer that would have been continuous with the high topography of both the Tharsis rise and the southern highlands. It is highly unlikely that this large regional aquifer was fully confined. Rather, it would have been locally confined from above by the cryosphere, but laterally continuous with unconfined aquifers in the Tharsis region and the southern highlands. Such a partially confined aquifer can be brought about by means other than topographic variations as well, as illustrated in Figure 6. In the absence of significant topographic relief (Figure 6a), locally confined aquifers that are laterally continuous with unconfined aquifers can be produced by variations in the thickness of the cryosphere (Figure 6b) as a result of lateral variations in either the thermal conductivity of the surface materials or the local heat flux. The same outcome could also be produced by variations in the height of the water table (Figure 6c), which could result, for example, from hydrothermal convection or a localized pressurization mechanism operating within or beneath the aquifer.

Thus the thickening cryosphere is not simply compressing a closed aquifer. Rather, as pressure is generated beneath the freezing front, it simultaneously diffuses away toward the edges of the confined portion of the aquifer. The actual pressure generated depends upon the rate at which the freezing front advances, the area over which the aquifer is confined, and the hydraulic properties of the aquifer. If either the confined portion of the aquifer is small or the thickening of the cryosphere is a slow process, the pressure should diffuse away essentially as fast as it is generated. Alternatively, if the confined portion of the aquifer is sufficiently large or the cryosphere thickens rapidly, significant pressures should be generated, possibly approaching those generated within a fully confined aquifer.

6.2. Pressurization of Aquifers: Model Description

In order to test the feasibility of the thickening of the cryosphere as a means of producing high pressures in Martian aquifers, we consider the hydrologic response to the cryospheric thickening for a number of climate change scenarios and aquifer geometries. The time dependant thickening of the cryosphere is modeled using a fully explicit finite difference code. The temperature evolution within the crust is modeled using the 1-D heat equation:

$$\frac{\partial T}{\partial t} = \frac{1}{\rho c} \frac{\partial}{\partial z} \left(k \frac{\partial T}{\partial z} \right),$$

where \(k\) is the thermal conductivity and \(c\) is the specific heat. The rate of change of the depth of the melting isotherm
(\(\tau_{\text{m}}\)) is calculated using the difference between the temperature gradient above and beneath the freezing front:

\[
\frac{\partial\tau_{\text{m}}}{\partial t} = \frac{1}{n \rho L} \left( k_1 \frac{\partial T}{\partial z} - k_2 \frac{\partial T}{\partial z} \right), \tag{28}
\]

where \(n\) is the porosity, \(L\) is the latent heat of fusion of water, \(\rho\) is the density of water, \(k_1\) is the thermal conductivity above the freezing front, and \(k_2\) is the thermal conductivity below the freezing front. For simplicity, we assume a fully ice-saturated cryosphere, though it is likely that the ice content of the pore space will decrease to zero toward the surface at a steady state ice table [Mellon et al., 1997]. The thermal conductivities are modeled as in Section 4.5.

[62] Once the thickness of the cryosphere as a function of time is computed, we model the pressure that would be generated beneath the growing cryosphere as a function of time for a given aquifer geometry. The pressure generated at the freezing front is found by considering a finite-radius aquifer of thickness \(D\) beneath the freezing front. As the freezing front moves an incremental distance \(\delta z\) into this aquifer, the expansion of the water as it freezes to form ice expels an amount of water equivalent to a layer of thickness \(n(\alpha_v - 1)\delta z\), where \(\alpha_v\), the coefficient of volumetric expansion of water upon freezing, is approximately 1.09, and \(n\) is the porosity. This water is forced into the aquifer beneath the freezing front, resulting in a pressure increase of \(\Delta P\):

\[
\Delta P = \frac{1}{\beta} \ln \left[ \frac{D - \delta z + n(n_v - 1)\delta z}{D - \delta z} \right]
\approx \frac{1}{\beta} \frac{n(n_v - 1)}{D - \delta z}, \tag{29}
\]

where \(\beta\) is the combined compressibility of the aquifer matrix and water.

[63] Since the freezing front moves slowly relative to the vertical diffusion of the excess hydraulic head, and since the vertical length scale of the aquifer is several orders of magnitude less than the horizontal, the aquifer properties are vertically averaged and diffusion is considered in the horizontal direction only (i.e., \(\partial h/\partial z = 0\)). The evolution of the hydraulic head produced by this excess pressure in the confined portions of the aquifer is calculated using a finite difference code representing the elastic response of the aquifer as governed by the consolidation equation (equation (5)) expressed in axisymmetric cylindrical coordinates with a pressure source term:

\[
\frac{\partial h}{\partial t} = \frac{1}{\rho_v g} \frac{\partial}{\partial r} \left[ K \cdot r \frac{\partial h}{\partial r} \right] + \frac{1}{\rho_v g} \frac{\partial (\rho P)}{\partial t}, \tag{30}
\]

where \((\partial P/\partial h)_{\text{cryos}}\) is the rate of pressurization at the freezing front from equation (29). In the unconfined portions of the aquifer, the net flow of water into a column of aquifer results in an increase in the hydraulic head through a direct increase in the water table height, rather than through a confined pressurization. It is also necessary to account for the fact that the aquifer thickness now varies with radius due to the increase in the water table height. If the hydraulic head is taken with respect to the base of the aquifer, such that it is equal to the unconfined aquifer thickness, then flow is modeled by

\[
\frac{\partial h}{\partial t} = \frac{1}{r \cdot n_{\text{aop}} \frac{\partial}{\partial r}} \left( K \cdot h \cdot r \frac{\partial h}{\partial r} \right). \tag{31}
\]

where \(n_{\text{aop}}\) is the porosity at the water table.

6.3. Pressurization of Aquifers: Model Results

[64] We consider instantaneous climate changes from 270 K to 220 K, and from 250 K to 220 K, representing a transition from the warm wet climate conditions on early Mars to the cold conditions of today in a single rapid event. We also consider more plausible gradual climate change scenarios, with the temperature changing from 270 K to 220 K over the course of 60 ka and 1 Ma. The former case corresponds to half of the period of an obliquity oscillation, while the latter case is a somewhat more conservative timescale for climate change. The resulting cryosphere thickness as a function of time for the four scenarios is shown in Figure 7. The cryosphere thickness is sensitive to the thickness of ice-free regolith above the permafrost, here assumed to be negligible, and the geothermal heat flow, here assumed to be 50 mW/m². This heat flux is appropriate for the Hesperian epoch on Mars [Hauck and Phillips, 2002], in which most of the outflow channels formed [Baker, 1982; Carr and Clow, 1981; Masursky et al., 1977].

[65] We use a generic aquifer geometry in which a circular region of the aquifer is confined from above by the thickening cryosphere, but the adjacent aquifer is unconfined. Our nominal model consists of an aquifer with a confined region radius (\(R\)) of 1000 km and an aquifer thickness (\(D\)) of 3 km. This aquifer thickness is roughly equal to the total change in cryosphere thickness during the simulation. We also consider confined regions with radii ranging from 500 to 3000 km, as well as initial aquifer thickness ranging from 1 to 5 km. The megaregolith aquifer model is representative of several of the outflow channel source regions to the east of Tharsis, while those further up on the Tharsis rise may be better represented by the basaltic aquifer model. Due to the difficulty constraining the hydraulic behavior of basaltic aquifers under varying conditions of lithostatic and fluid pore pressure, we limit the following investigation to the megaregolith aquifer model. Lower pore pressures would be expected for basaltic aquifers due to their higher permeability relative to megaregolith aquifers, thus the pore pressures generated in our models are likely upper limits.

[66] Figure 8 plots the aquifer pore pressure at the base of the cryosphere in the center of the confined region as a function of time for the aquifer geometries and climate change scenarios described above. Also included in the figure is the lithostatic pressure at the base of the cryosphere, which increases with time as the cryosphere thickens. For all scenarios, the pore pressure rises steeply with time during the early period of rapid cryosphere thickening. As the rate of propagation of the freezing front slows, the pore pressures plateau out. After the freezing front passes through the base of the megaregolith, a smaller volume of water is expelled from the freezing front per unit increase in cryosphere thickness due to the decrease in porosity. As a
the lithostatic pressure. For the change in surface temperature from 270 to 220 K over the course of 60 ka, the pore pressure for the 3000 km radius confined aquifer reaches, but does not exceed, the lithostatic pressure (Figures 8c and 8d). Smaller confined aquifers result in lower pore pressures. For the change in surface temperature from 270 K to 220 K over the course of 1 Ma, the pore pressure remains well below the lithostatic pressure for all confined aquifer sizes (Figure 8e). During these slower climate changes, the pore pressure has ample time to diffuse away from the slowly advancing freezing front toward the unconfined portions of the aquifer. For an instantaneous change from 250 K to 220 K, the greater initial cryosphere thickness results in a greater thermal diffusion timescale at the base of the cryosphere and a slower initial rate of cryospheric thickening, leading to the generation of significantly lower pressures within the aquifer (Figure 8f). These more plausible scenarios produce pore pressures significantly less than the lithostatic pressure at all times.

For the instantaneous climate change from 270 K to 220 K, most scenarios result in maximum pore pressures approaching, but not exceeding, the lithostatic pressure (Figures 8a and 8b). As the fluid pore pressure approaches the lithostatic pressure, the fracture apertures rapidly increase (equation (16)), resulting in a sharp rise of the permeability (equation (12)). This allows for a more rapid lateral diffusion of the pressure away from the confined portion of the aquifer, producing a negative feedback cycle that prevents fluid pressures from exceeding the lithostatic pressure. For the confined aquifers with radii up to 2000 km, the pore pressures approach the lithostatic pressure during the early period of rapid cryospheric thickening, before plateauing at lower values as the rate of thickening decreases. For the locally confined aquifer with a radius of 3000 km, pore pressures exceed the lithostatic pressure during this early period by approximately 0.4 MPa. However, as the rate of thickening of the cryosphere decreases, the pressure diffuses away more rapidly than it builds up and the aquifer quickly returns to sublithostatic pore pressures.

For more plausible climate change scenarios, the pore pressure at the base of the cryosphere never exceeds the lithostatic pressure. For the change in surface temperature from 270 to 220 K over the course of 60 ka, the pore pressure for the 3000 km radius confined aquifer reaches, but does not exceed, the lithostatic pressure (Figures 8c and 8d). Smaller confined aquifers result in lower pore pressures. For the change in surface temperature from 270 K to 220 K over the course of 1 Ma, the pore pressure remains well below the lithostatic pressure for all confined aquifer sizes (Figure 8e). During these slower climate changes, the pore pressure has ample time to diffuse away from the slowly advancing freezing front toward the unconfined portions of the aquifer. For an instantaneous change from 250 K to 220 K, the greater initial cryosphere thickness results in a greater thermal diffusion timescale at the base of the cryosphere and a slower initial rate of cryospheric thickening, leading to the generation of significantly lower pressures within the aquifer (Figure 8f). These more plausible scenarios produce pore pressures significantly less than the lithostatic pressure at all times.

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Figure 8
adjust to the changing fluid pore pressure (dotted line in Figures 8a and 8b). In this case, the pore pressure rises much more rapidly and exceeds the lithostatic pressure by a substantial margin. Similarly large pressures could be achieved in an unfractured sedimentary aquifer of sufficient vertical and horizontal extent, however there is no evidence that such an aquifer exists in the region of the outflow channel source regions.

[71] Similar simulations have been run investigating the climate change scenarios and aquifer geometries described above over a wider range of parameter space in the hydrologic model. Model results are most sensitive to the uncertainty in the fracture aperture, due to the cubic dependence of the permeability on this parameter. Using an uncompressed fracture aperture of $10^{-4}$ m, at the lowest end of the plausible range, super-lithostatic pore pressures can be generated for several of the model scenarios, including the temperature changes from 270 to 220 K both instantaneously and over 60 ka for confined aquifer radii greater than 1000 km. However, the high discharges inferred for the outflow channel floods are suggestive of higher than average permeability, rather than lower.

[72] The circum-Chryse region, in which most of the outflow channels occur, is similar in size to the 2000 km radius confined aquifer modeled above. Thus it would seem that if the aquifer in the region were confined over an area somewhat greater than the area of the Chryse basin (e.g., over a 3000 km radius area), it would be possible to produce super-lithostatic pore pressures during an instantaneous change in surface temperature from 270 K to 220 K. However, the model pore pressures peak in the center of the confined region and decrease toward the periphery, while the outflow channel source regions are all along the margin of the circum-Chryse region. Thus for such a hypothetical confined aquifer in this region, the super-lithostatic pore pressures would be generated in the center of the Chryse basin, and not where the outflow channel source regions are located.

[73] If the effects of topographic variations were to be considered, then an extra component of hydrostatic pressure would be added to the pore pressure in the aquifer beneath topographic lows. However, many of the source regions of the circum-Chryse outflow channels are relatively high up on the Tharsis bulge and show no evidence for a preferential occurrence in local topographic lows. If the aquifer pressurization mechanism acted over the entire circum-Chryse region, the maximum hydrostatic component of the pressure would occur in the topographically lower center of the Chryse basin, enhancing the higher pressures being produced there. Thus the location of the outflow source regions at higher elevations on the margins of the Chryse basin suggests a more localized pressurization mechanism focused at the individual source regions.

[74] A further argument against a climatic mechanism as the driving force behind the outflow channel floods lies in the observations that the channels have ages ranging from Hesperian to early Amazonian, a time period spanning roughly 1 billion years [Carr and Clow, 1981; Tanaka and Skinner, 2004], and that several of the outflow channels show evidence for multiple episodes of flow [Hanna and Phillips, 2003; Williams et al., 2000]. Thus it would be necessary to have repeated near-instantaneous climate changes from 270 to 220 K recurring over a period of roughly 1 Gya to explain the temporal distribution of outflow channel events. While a number of workers have argued for the importance of episodic climate changes on Mars [Baker et al., 1991; Head et al., 2003], temperature swings as large and rapid as are required to form the outflow channels through this mechanism seem unlikely. Mechanisms commonly invoked for secular climate change, such as erosion of the atmosphere by impacts and solar wind stripping [Brain and Jakosky, 1998] act over longer timescales. Gillick et al. [1997] found that the injection of 1 to 2 bars of CO$_2$ into the Martian atmosphere resulted in a temperature increase to only 250 K, however the subsequent cooling took place over hundreds of Ma. Rapid climate changes attributed to obliquity oscillations will be limited by the CO$_2$ content of the polar caps, which has been shown to be on the order of tens to hundreds of mbar [Mellon, 1996], much less than that required for a strong greenhouse effect.

[75] In summary, the difficulty with producing super-lithostatic pore pressures through a climate change lies both in the slow diffusion of the thermal wave into the crust and in the negative feedback cycle between the pore pressure and the permeability. This effect would be magnified for basaltic aquifers, which have a significantly higher permeability than that in the megaregolith model. These results have implications for any mechanism of aquifer pressurization that relies upon processes acting over timescales of tens of thousands to millions of years. The work presented here suggests that a more rapid, repeatable, and localized mechanism for pressure generation would be favored as a driving force behind the formation of the outflow channels, such as a magmatic intrusion into the aquifer or a tectonic event.

[76] Due to the unconstrained nature of the problem, it is of course not possible to fully prove or disprove the theory that the outflow channels were generated by the pressurization of an aquifer beneath a downward propagating freezing front. However, this study sheds some light on the details and problems of this theory. It does not appear possible to produce super-lithostatic pore pressures solely through the freezing of a megaregolith-type aquifer for plausible climate change scenarios and aquifer geometries. This does not rule out the possibility that this pressurization mechanism could have produced the requisite pressures in concert with another mechanism. For example, this mechanism may have played a secondary role by causing a general increase in aquifer pore pressures to sublithostatic levels over a large region, thereby allowing a more localized and rapid pressurization mechanism to push pore pressures to super-lithostatic levels at the outflow channel sources.

7. Conclusions

[77] We have presented here a generalized hydrologic model of the Martian crust based on a combination of terrestrial and lunar analogs, in which the relevant hydraulic properties are represented in a physically realistic and self-consistent manner. Much of the hydrologically active crust of Mars can be represented by the megaregolith aquifer model, based on the inferred effects of impacts on the hydrologic properties of the crust. Younger terrains are better represented as either basalts, such as the topmost
layers in the Tharsis rise, or sediments, such as the surface materials within craters and elsewhere [Malin and Edgett, 2000b]. The hydrologic properties of all of these crustal materials will vary with both depth and pore pressure.

[75] The sensitivity of the hydraulic properties to the pore pressure was highlighted in the study of the pressurization of a partially confined aquifer beneath a thickening cryosphere. We found that even for an unrealistically rapid climate change, the diffusion of the excess pressure toward the unconfined portions of the aquifer precluded the attainment of super-lithostatic pore pressures for almost all cases considered. This lateral diffusion was accentuated by the increase in permeability at high pore pressures, resulting in a negative feedback cycle between pore pressure and permeability. This feedback will be important for any hydrologic process involving large and widely varying pore pressures.

[79] Hydrological modeling of groundwater processes on Mars is essential for understanding the many observed water-related features on the surface. While the nature and hydraulic properties of the materials in the subsurface are poorly constrained, the model developed here provides a baseline with which we can begin to quantitatively test current hypotheses regarding Martian hydrological processes, as has been done here for the origin of the outflow channels. The parameters contained in this study are best estimates based on our current understanding, but will be subject to revision as more data become available. The forthcoming data from the SHARAD and MARSIS radar sounders will likely shed new light on the distribution and abundance of Martian groundwater, thereby providing a much-needed constraint on the aquifer model.

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